

MASS-BALANCE MEASUREMENTS IN ALASKA AND SUGGESTIONS FOR SIMPLIFIED OBSERVATION PROGRAMS

BY

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Trabant, D.C., and March, R.S., 1999: Mass-balance measurements in Alaska and suggestions for simplified observations programs. *Geogr. Ann.*, 81 A (4): 777–789.

ABSTRACT. US Geological Survey glacier fieldwork in Alaska includes repetitious measurements, corrections for leaning or bending stakes, an ability to reliably measure seasonal snow as deep as 10 m, absolute identification of summer surfaces in the accumulation area, and annual evaluation of internal accumulation, internal ablation, and glacier-thickness changes.

Prescribed field measurement and note-taking techniques help eliminate field errors and expedite the interpretative process. In the office, field notes are transferred to computerized spreadsheets for analysis, release on the World Wide Web, and archival storage. The spreadsheets have error traps to help eliminate note-taking and transcription errors. Rigorous error analysis ends when mass-balance measurements are extrapolated and integrated with area to determine glacier and basin mass balances. Unassessable errors in the glacier and basin mass-balance data reduce the value of the data set for correlations with climate change indices.

The minimum glacier mass-balance program has at least three measurement sites on a glacier and the measurements must include the seasonal components of mass balance as well as the annual balance.

Introduction

The US Geological Survey (USGS) began mass balance measurements on Gulkana and Wolverine Glaciers in Alaska (Fig. 1) during 1966 as part of the United States' contribution to the International Hydrologic Decade study of glaciers. In addition to the mass-balance measurements, runoff and meteorological recorders were installed in the basins. Since 1966, the net balance, accumulation, ablation, accumulation area ratio (AAR), and equilibrium line altitude (ELA) for Gulkana and Wolverine Glaciers have been published by the World Glacier Monitoring Service (Kasser 1967; Müller 1977; Haeberli 1985; Haeberli and Müller 1988; Haeberli and Hoelzle 1993; and Haeberli *et al.* 1998). Mass balance data are also regularly reported in the *Glacier Mass Balance Bulletin* (for example, Haeberli and Herren 1991). Detailed results from 1966 and 1967 were reported by Meier *et al.*

(1971) and Tangborn *et al.* (1977), respectively. Measured winter snow balances and annual balances from 1966 to 1977 were reported by Meier *et al.* (1980). Balance studies were relatively intense until the mid-1970s, after which spatial sampling was reduced to three sites used as indices for mass balance. During the early 1970s, measurements at the three index sites were expanded to include ice-motion and surface-altitude change (for determining glacier-volume change). The current program continues mass-balance measurements at three sites on each glacier (Fig. 1). The current measurements at Gulkana and Wolverine Glaciers are intended to sustain them as parts of the USGS Benchmark Glacier Network of the United States (Fountain *et al.* 1997, p. 8). The purpose of this paper is to describe the essential components of the mass-balance measurement program operated in Alaska by the USGS and to use the 30-year time series of measurements and climate regime shifts to design a minimum program for mass balance measurements on other glaciers.

Gulkana Glacier

Gulkana Glacier (63°16'N, 145°28'W, Fig. 1) is a south-facing branched valley glacier, 8.5 km long and 19.3 km² in area, on the southern flank of the eastern Alaska Range in interior Alaska. The local climate is continental not only because it is 310 km from the nearest sea water, but also because it is shielded from moisture-laden storms by two orographic barriers, the Chugach Mountains and either the Wrangell or Talkeetna Mountains, depending upon the storm direction. The average ELA is about 1770 m. Gulkana Glacier is about 2 km from the end of a seasonally passable road; therefore, summer access by hiking and backpacking equipment is possible. Springtime access is normally by snowmobile, beginning from the plowed highway about 8 km

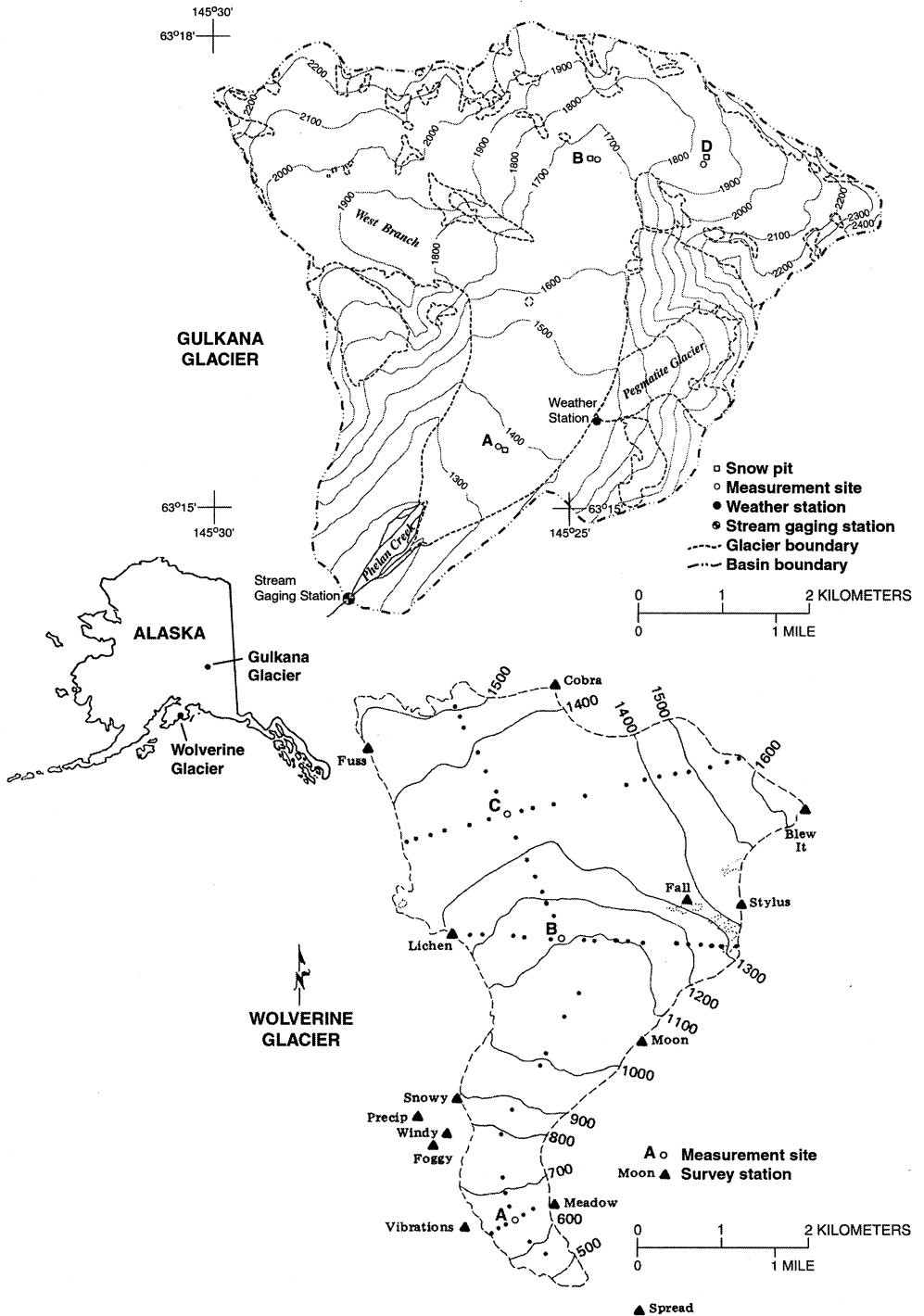


Fig. 1. Location map for the US Geological Survey benchmark glaciers in Alaska. On Wolverine Glacier, geodetically surveyed points along the longitudinal and transverse profiles are shown as filled dots. The mass-balance measurement sites on both glaciers are shown as open circles.

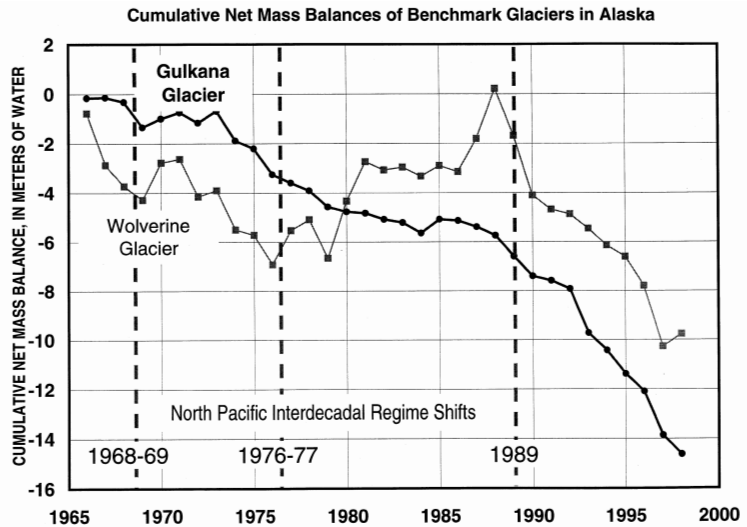


Fig. 2. Cumulative net mass balance for Gulkana and Wolverine Glaciers, Alaska, 1966 to 1998. Vertical lines mark recognized interdecadal regime shifts of the climate of the northern Pacific basin (Overland *et al.* 1999). Note that the mass of Wolverine Glacier increased between the 1976–77 and 1989 regime shifts. The rate of mass loss of Gulkana Glacier was somewhat reduced during the same period.

from the terminus. Fall access is usually by helicopter. The helicopter-supported trip is used to resupply the equipment cache and cabin. Index-site glacier-surface and summer-surface altitudes, measured winter balance, and net firn and ice balance from 1975 to 1983 were reported by Mayo and Trabant (1986). Data for 1992 and 1993 were published by March and Trabant (1996, 1997) and for 1994 by March (1998). The cumulative mass balance of Gulkana Glacier (Fig. 2) indicates a net thinning of 14.6 m (water equivalent) since 1966. The rate of thinning has not been constant (Fig. 2). Furthermore, the mass balances of Gulkana Glacier have not yet been checked for systematic errors by comparison with an independently measured glacier-volume change. The error in each year's measurement has been estimated to be about 0.2 m (Meier *et al.* 1971; Tangborn *et al.* 1977). Assuming that each year's errors are truly random standard errors, the cumulative standard error for the 32-year period of record is 1.1 m.

Wolverine Glacier

Wolverine Glacier (60°24'N, 148°54'W, Fig. 1) is a south-facing valley glacier, 7.1 km long and 18.0 km² in area, in the Kenai Mountains of the Kenai Peninsula in south-central Alaska. Because Wolverine Glacier is only 5 km from saltwater near the southern coast of Alaska, the local climate is maritime. The average ELA is about 1200 m. Because Wolverine Glacier is in a remote area, access is al-

ways by helicopter. Annual, summer, and winter balances from 1966 to 1982 were correlated with temperature trends by Mayo and Trabant (1984). A glacier volume change for the period 1974 and 1985 was estimated on the basis of geodetically surveyed profiles (Mayo *et al.* 1985). The cumulative net mass balance of Wolverine Glacier was adjusted to agree with measured volume change. The correction decreased the annual area-averaged net-mass-balance by 0.46 m water equivalent. Applying this adjusted-area integration scheme to the entire record resulted in a cumulative mass balance for Wolverine Glacier (Fig. 2) that indicates a net thinning of 9.8 m (water equivalent) since 1966. The error in each year's measurement has been estimated to be about 0.2 m (Meier *et al.* 1971; Tangborn *et al.* 1977); therefore the cumulative error for the 32-year period of record is estimated to be 1.1 m. The cumulative mass balance of Wolverine Glacier (Fig. 2) shows a growth period between 1976 and 1988 followed by an unbroken series of negative mass balances until 1998, constituting the most rapid rate of mass loss since the measurements began.

Other glaciers

The operational techniques developed for measuring the mass balance of Gulkana and Wolverine Glaciers have been used on other glaciers in Alaska. Notably, a 23-year record for 10 sites on Black

Rapids Glacier has been reported by Heinrichs *et al.* (1995). Black Rapids Glacier is a 40 km long surge-type glacier located in the central Alaska Range, about 35 km northwest of Gulkana Glacier. Black Rapids Glacier has a surface area of 246 km². During an intensive year of measurements on Columbia Glacier (Mayo *et al.* 1979), 60 sites were measured on the 68-km long, 1100 km² glacier, which calves into Prince William Sound near Valdez, Alaska. Four years of mass balances have been reported for McCall Glacier, where as many as 57 measurement sites were maintained (Trabant and Benson 1986). McCall Glacier is a north-facing valley glacier on the north flank of the Brooks Range in the northern part of Alaska (69°18'N, 143°48'W). It is about 8 km long and has a surface area of 7.2 km². Other Alaska glaciers that are less well reported and have shorter periods of observations are Knik Glacier (Mayo and Trabant 1982) near Palmer, and Kahiltna and Traleika Glaciers in Denali National Park.

Field methods

Benchmark glaciers

Mass-balance measurements on Gulkana and Wolverine Glaciers are normally made three times each year by a field party of two. The data are recorded and reported using the combined mass-balance system defined by Mayo *et al.* (1972) with the addition of evaluations of internal accumulation (Trabant and Mayo 1985) and internal ablation (March and Trabant 1997).

Measuring the surface mass balance of any glacier is simple in concept; however, commonly overlooked details in fieldwork and record keeping commonly complicate data analysis. Field techniques have been developed to both expedite and standardize fieldwork. The USGS glacier field program in Alaska includes (1) prescribed standards for recording notes; (2) redundant measurements; (3) measurements for analyzing and correcting for leaning, bending, or sinking of stakes; (4) reliable measurement of seasonal snow as deep as 10 m; (5) absolute identification of summer surfaces in the accumulation area; and (6) annual evaluation of internal accumulation.

Before the field trip begins, an estimate of the current location of each stake is plotted on a large-scale map. These maps include a history of displacements, changes in the length of the stake, and mass-balance readings for each stake near an 'index point'. An index point is a fixed horizontal-

space coordinate that is the focus for all of the nearby measurements. Stakes are installed about one year's displacement upglacier from the index point and are replaced at approximately one year intervals. Glacier motion moves the stake toward the index point, and usually beyond it before a stake is abandoned.

Carefully recorded field measurements are a fundamental necessity. Field measurements are recorded on water-resistant paper, in a strictly serial chronology of the fieldwork, and include all of the field-measured components that constitute each value (examples below).

When new stakes are installed, embossed-tape labels are attached to the top of each stake section (typically of 3 m length). Each label contains the year, location on the glacier, and the distance, in meters, from the bottom of the stake to the top of each section. For example, the bottom section of a stake installed during 1998 near index site B would be labeled 98-B-3m. The 3-m-long stake section above that would be labeled 98-B-6m.

Glacier stakes are 1 inch inside diameter, thin-wall steel tubing, cut to 3 m lengths with holes drilled at 0.5 m intervals along the length. The holes allow water to be exchanged and serve as auxiliary length-reference points. Stake connectors are 0.3 m lengths of heavy-duty tubing that will freely slide inside the stake sections. A snap-ring, installed in a machined slot near the center of the connector, prevents the connector from falling completely into the lower stake. The bottom of the stake is plugged with a tightly fitting wooden dowel that extends a few centimeters beyond the end of the metal stake.

Measurement of the glacier surface at a stake is made relative to the bottom of the stake; that is, the recorded stake height of the glacier surface is the length of stake that is below the surface. At the same time, the surface material is identified and recorded. For example, if the top of a 6 m stake is 0.53 m above the glacier surface and the surface material is snow, the notebook entry is '6.00 - 0.53m snow top'. The stake height of the surface is intended to represent the average surface and ignores ablation pits or minor drifting caused by the stake itself. The recognized surface materials are snow, firn, superimposed ice, and ice (Mayo *et al.* 1972, p. 3-14). If the surface material is discontinuous (for example, fresh snow in the ablation zone), the percentage of cover within a few tens of meters of the stake is noted. If the surface material is snow or superimposed ice, its average depth and density are

determined and recorded. The fundamental measurement of the stake height of the glacier surface is so important that it is independently measured during the geodetic survey that determines the location of the stake.

All stakes and the local glacier-surface slope and altitude are geodetically surveyed during each visit using a theodolite and laser distance-measuring equipment. The surveying data are usually gathered to support a combination of resection and foresight calculations (Mayo and Trabant 1982, p. 3–5). Stake location, lean of the stake, and the stake height of the glacier surface are calculated from the survey data. Stake locations are calculated for the bottom of the stake, and therefore represent locations of a parcel of glacier ice, assuming that the bottom of the stake is fixed in the ice. Possible sinking or rising of stakes is analyzed in the office by comparing successive stake heights of unchanged stratigraphic horizons (March 1998, p. 16–19). Stake lean is determined by surveying two places on the stake separated by about 1.5 to 2.0 m. One of the surveying targets is always chosen to be the average surface of the glacier on the stake. Surveying this point produces a redundant determination of the stake height of the glacier surface.

Surveyed targets must be carefully identified in the field notes. For example, the two targets on a single stake will be identified by the same stake name, followed by the stake height of the target. For example, if a surveying target is the 6.5 m hole in a stake that was installed during 1998 near index A, the notebook entry for the target will be designated 98-A-6.5m. The glacier-surface height on the stake is always specified as b' . Therefore, the notebook entry for that target would be 98-A- b' .

During the fall visit to the accumulation area, stratigraphic markers are placed on the summer surface. A 1.5 m square of plywood with a hole in the center is placed over some of the stakes. Sometimes sawdust is spread in a recorded direction and distance from a stake. Spring snow depth and year-end firn accumulation are determined by coring or steam drilling as deep as the plywood, or by digging a snow pit or coring to the sawdust. After the summer surface is unambiguously identified, snow probing can be used to determine the average depth for the area near the index site. However, snow probing, especially in the accumulation area, is not always reliable and must be verified by some direct observation. Steam drilling to plywood is a quick way to measure seasonal snow as deep as 10 m.

Snow and firn densities are commonly measured

in snow pits where a continuous column of snow is sampled as the pit is dug. Core sampling for density is generally less accurate than sampling in a pit wall because some snow is invariably lost and the depth of the retrieved sample can be questionable. Continuous-penetration snow-core samplers work well when snow depths are less than a few meters, but tend to compress low-density layers deep within the snow, which can make it difficult to be sure that the entire snow thickness has been sampled. When the snow is more than about 5 m deep, continuous-penetration samplers are difficult to drive through the entire depth.

The temperature of the snow-firn interface is measured in snow pits or cores during the spring visit. This interface temperature is used for estimating the amount of internal accumulation that will occur (Trabant and Mayo 1985, p. 113–117). Assessment of internal accumulation is a routine part of the mass-balance analysis of glaciers in Alaska (March 1998, p. 19).

Before leaving a measurement site, the final notebook entries record the total length of all stakes and the type and location of stratigraphic markers. Stake lengths are usually adjusted during each visit. During fall or early winter visits, stakes are extended; during spring and early summer visits, stakes are usually left shorter than they were when found.

Large and high-latitude glaciers

On the largest glaciers in Alaska and those at the highest latitudes, the traditional definitions of firn and summer surface, which are fundamental to the stratigraphic interpretation of net balances, become ill-defined. The definitions begin to fail in the upper parts of Benson's (1959, p. 25; see also Cofin *et al.* 1990, p. 220) percolation facies and completely fail above the dry-snow line in the dry-snow facies. In the mountains near the southern gulf coast of Alaska, the separation between the percolation and dry-snow facies (the dry-snow line) occurs at about 3000 m altitude. Many of the large glaciers near the southern gulf coast of Alaska reach 3500 to 4000 m altitude and have large parts of their accumulation areas near or above the dry-snow line (for example, Columbia Glacier; Trabant and Mayo 1985, p. 116). The altitude of the dry-snow line decreases with latitude (see Benson 1967, p. 1048) so the accumulation areas of most of the glaciers in the Brooks Range of northern Alaska are near or above the dry-snow line. Where

accumulation areas are near or above the dry-snow line, 'summer' melt seldom wets more than the upper few millimeters of snow, and melt surfaces are repeatedly buried by fresh snowfalls. In these areas, some 'firn' layers (snow that has survived some 'summer' melt) may be only a few days old and multiple 'summer surfaces' are formed each year. Depth hoar, the natural stratigraphic marker of Greenland and Antarctica, is formed in the Brooks Range glaciers and can commonly be used to identify summer surfaces. However, in the relatively warm, high-precipitation maritime climate near the gulf coast of Alaska, depth hoar rarely develops in the snow pack. On these glaciers, a dated artificial, or unique, stratigraphic marker must be in place and must be recovered in order to define the net mass balance.

Passive stratigraphic markers such as sawdust, plywood, dyes, and deeply colored, powdered fruit drink concentrates have been successfully used on the benchmark glaciers as well as on glaciers studied for special projects. The passive marker is placed on the surface in the accumulation area at a specified location, usually near a balance stake, and serves as an indisputable, absolutely dateable, stratigraphic marker. However, where annual accumulations approach 5 m depth (for example, on Columbia Glacier; Trabant and Krimmel 1997) balance stakes are buried or do not survive and the location of the passive marker may be lost.

Active markers are used to help recover the locations of passive stratigraphic markers on large glaciers with heavy snow accumulations. Both radio beacon transmitters and strong magnets have been used. The radio beacons, originally designed for tracking large animals and fish, were modified by increasing the length of the antenna and the size of the battery. These modifications extended their range, especially when buried by snow, and their operational life. A receiver with a directional antenna and field-strength analyzer can be used to quickly locate transmitters with enough accuracy so that a snow pit or core will encounter a stratigraphic mark that is 10 m in diameter. Another 'active' marker is strong permanent magnets left at the site of the passive marker. Magnets can be located by mapping the magnetic anomaly with a magnetometer that has a sensitivity of a few gammas (Harrison *et al.* 1978). An additional benefit of using magnets is that the depth to the magnets can be calculated from the geometry of the field strength measurements. On Columbia Glacier, the technique was verified by digging to and re-

covering magnets after one year's accumulation. On the other hand, mapping the magnetic anomaly can be completely frustrated if the Earth's magnetic field is rapidly fluctuating. Furthermore, the magnetic anomaly may be relatively small, especially if the magnets are deeply buried. This means that the magnetic search must begin near the magnets or they will never be found. Both radio beacons and magnets were used on Columbia Glacier during the study period (Mayo *et al.* 1979), but were not mentioned as field techniques. Radio beacons and magnets have been successfully located for periods as long as 3 years and at depths as deep as 15 m.

Occasionally, natural stratigraphic markers are helpful. The August 18, 1992 eruption of a flank peak on Mt Spurr Volcano deposited ash on Columbia Glacier. That ash was easily identified in crevasses throughout the accumulation area of Columbia Glacier during August 1997. Measurements to this absolutely dateable ash horizon validated interpretation of the intervening stratigraphy at nine sites, spanning 1500 m altitude in the accumulation area of Columbia Glacier (Trabant and Krimmel 1997).

Office procedures

In the office, field notes are transferred to computerized spreadsheets for analysis, release on the World Wide Web (<http://ak.water.usgs.gov/glaciology/>), and archival storage. The spreadsheets have error traps to help eliminate field-note blunders and transcription errors. Computerized spreadsheets also promote consistent analysis. This consistency is especially important when several people are involved in the collection and analysis of mass-balance data. Directly measured data are given higher priority than estimated values in the spreadsheet-based analysis.

The surveying and mass-balance measurements are closely integrated in the spreadsheet analysis of mass balance and glacier motion. This combined analysis allows correcting balance readings for the effects of stake deformation (lean, bend, or bow) before mass balance calculations are begun. For example, stake lean is mathematically corrected by using the two surveyed points on the stake and by assuming that the base of the stake has not moved relative to the surrounding ice. A corrected stake height of the glacier surface (that is, a corrected mass-balance reading) on a theoretical vertical stake is calculated using the locally measured sur-

face slope of the glacier (Mayo and Trabant 1982, p. 7–9). If stake slip or rise, relative to the surrounding ice, is detected, the slip or rise effect is also removed before mass balance calculations begin. An integral part of the spreadsheet analysis is a history of the bends in each stake. The bend history is used when determining the three-dimensional location of the base of the stake and for correcting the stake height of the glacier surface for mass-balance readings. A ‘bow’ is a deformation that is within the elastic limits of the stake; that is, a deformation from which the stake will recover on its own. Bow solutions are usually applied when spring snowpacks deform stakes that are later found to be straight after the snowpack melts.

Point mass balances

Because mass-balance field measurements are made relative to time-transgressive stratigraphic horizons (summer surfaces), adjustments must be made each year to determine the maximum winter balance, net balance, and fixed-date annual balance. Furthermore, field visits seldom coincide with the exact beginning and end of the stratigraphic (net-balance) or hydrologic (fixed-date or annual-balance) year. Indeed, in Alaska, on all but the smallest glaciers, it is rare that the stratigraphic year ends simultaneously over the entire glacier. Even on relatively small glaciers, like Gulkana and Wolverine, it is much more common for new snow to begin accumulating on the upper parts of a glacier several weeks before ablation ends at the lowest altitudes. This results in a time-transgressive summer surface (stratigraphic horizon) that represents ‘the end of the net-balance year’. Choosing a single, glacier-average date for the end of the stratigraphic year and adjusting field measurements to the glacier-average date is done with a simple temporal-balance model.

A simple temporal-balance model is used to make the temporal interpolations that are necessary for evaluating the index-site balances: maximum winter balance, net balance, and annual balance. Modeled extrapolations are never published; data releases are always delayed until a bounding measurement has been made that will support interpolation with the temporal-balance model. The two-parameter linear model relates measured air temperature and precipitation to the measured mass-balance change at each index site (March 1998, p. 16–20). Temperature is lapsed from the recorder altitude using the wet-adiabatic lapse rate of

–0.66°C per 100 m altitude increase. The model estimates glacier ablation at the rate of about 3.5 to 5 mm water equivalent per degree Celsius above 0°C per day when the surface is snow, and twice that when the surface is ice or old firn. Glacier accumulation is estimated by the model to be 1.5 to 4.0 times the precipitation-gage catch when the lapsed temperature at the site is below 1.8°C. Melt (ablation) rates and the precipitation-catch multipliers are not fixed. Unique values are determined for each measurement period at each index site, so the modeled balances always agree with the measured balances. Thus, the model only temporally distributes the measured balance change at each site. The temporal corrections tend to be small; nevertheless, they reduce the uncertainty of the balance values assigned to each measurement site.

In an attempt to minimize errors and eliminate blunders, the stake height of the previous summer surface is calculated from the snow depth and balance reading for every visit. Barring stake slip or firn compaction, these assessments of the stake height of the summer surface should agree. They seldom do agree, however, and the generally random nature of the variation suggests that sampling error is the cause. In the spreadsheet analytical scheme, the measurements are combined using a weighted average. The weight for each measurement is proportional to the number of observations that were combined in the measurement. Making redundant field observations and combining them helps to eliminate field-note blunders. Furthermore, increasing the number of independent observations increases confidence in the value of the stake height of the summer surface. Redundant measurements are well worth the field time necessary to make them.

Winter balance. Mayo *et al.* (1972) defined three ‘winter’ balances in the stratigraphic system: the measured winter snow balance, the maximum winter snow balance, and the winter balance. The World Glacier Monitoring Service (WGMS) publishes a ‘winter’ balances that is not clearly any of the ‘winter’ balances defined by Mayo *et al.* (1972).

The measured winter snow balance (Mayo *et al.* 1972) is a late winter or spring measurement of the snow that is above the time-transgressive summer surface. The measurement is made in a short period of time and is intended to document snow depths while it is near its seasonal maximum throughout the glacier area. The measurement is usually made

shortly before melting begins on the lowest part of the glacier. The measured winter snow balance is often reported as the 'winter' balance. However, it is not a verified maximum balance.

The maximum winter snow balance of Mayo *et al.* (1972) is defined as the maximum snow mass during the balance year. On glaciers in Alaska, the instantaneous maximum snow balance usually occurs after there has been substantial melt of the snow (and sometimes glacier ice) at low altitudes on the glacier. The maximum winter snow balance occurs later in the spring on glaciers that have large altitude ranges, because the accumulation season lasts longer with altitude and the area of the lowest parts of the glaciers are relatively small. Therefore, the relatively small mass of snow lost low on the glacier is easily compensated by accumulation in the largest areas high on the glacier. The maximum winter snow balance is rarely directly measured and even if it was, a temporal-balance model would be necessary to verify it. After the time of occurrence of the maximum winter snow balance has been determined by temporal-balance modeling, the balance change between the time of the measured winter snow balance and the time of the glacier-averaged maximum winter snow balance is estimated for each measurement site using the temporal-balance model.

The winter balance is defined by Mayo *et al.* (1972) as the maximum balance relative to the balance at the time of the previous glacier-averaged seasonal minimum balance. The time of the glacier-average minimum was modeled because the time-transgressive seasonal minimum is seldom measured and never confirmed without modeled verification. The same is true for defining the time and magnitude of the glacier-averaged maximum.

Internal accumulation. Internal accumulation is the combination of two mechanisms by which water is permanently captured by cold firn. The first component is the amount of water frozen immediately when water percolates into cold firn and freezes. Surprisingly, this widely recognized component is the smaller of the two. The larger component is the amount of water that is retained as surface and capillary liquids. The capillary water will be frozen during the following winter. The released heat of fusion will help sustain strong temperature gradients across the thin fall snow cover. A strong temperature gradient promotes depth hoar development which forms the natural stratigraphic marker. The annual quantity of internal accumula-

tion is estimated using the annual minimum temperature of the snow-firn interface as defined by Trabant and Mayo (1985).

Errors. In an attempt to reduce errors, measurement sites are treated as small areas 25–75 m in radius, over which samples are taken and the balance is averaged. The area is chosen to be large enough, and enough samples are taken in the area so that when a quantity, such as snow depth, is averaged, the error caused by the glacier-surface and summer-surface roughness (up to several meters on some glaciers) is small. Nevertheless, errors arise from a combination of the errors in stake measurements, snow and firn densities, snow depth, and quantifying internal accumulation and ablation. Most of the sampling error must be estimated because the number of independent samples is insufficient to warrant rigorous error analysis. For example, seldom are there more than two balance stakes or more than one snow-density pit at a measurement site. Applying corrections for leaning, bent, and bowed stakes helps reduce the measurement bias and errors. Furthermore, the redundant measurements are known to have eliminated blunders that would have otherwise been incorporated into the data set but, again, these defy rigorous error analysis. The standard errors of rigorously analyzed individual elements of the mass balance are included in the spreadsheet output (for example, see March 1998, p. 17). The final site balances are assigned an error of ± 0.2 m water equivalent on the basis of how the currently used area-weighting factors reproduce the balances for the most intensively sampled and mapped balance analysis years of 1966 and 1967.

Glacier and basin mass balance

Glacier and basin mass balances are calculated from the measured site data by common area-integration techniques (for example, March and Trabant 1996, p. 15–22). Each site balance is assumed to be the average balance for some part of the glacier. The percentages change as the site altitudes and the glacier area–altitude distribution change with time.

Internal ablation. The glacier-average internal ablation is calculated by combining the effect of three internal and sub-glacial energy sources: geothermal heat, potential energy lost because of ice motion, and potential energy lost by water flowing

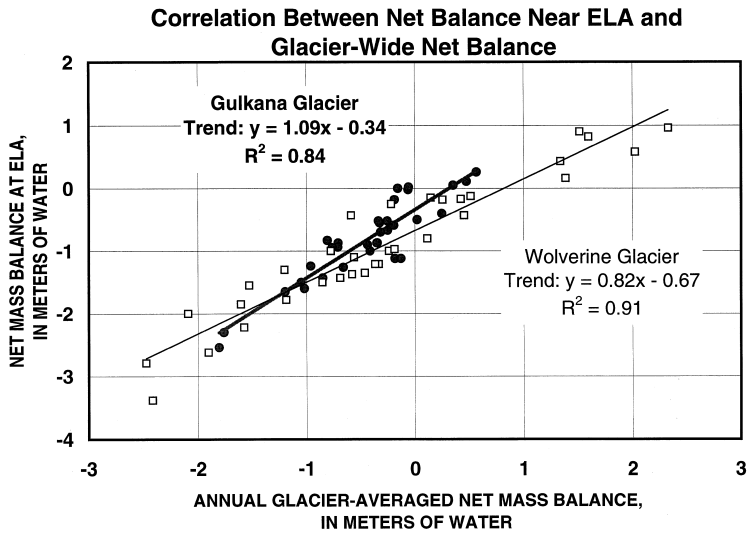


Fig. 3. Correlation between the net mass balances measured at sites near the Equilibrium Line Altitude (ELA) and the glacier-wide mass balances for Gulkana and Wolverine Glaciers, Alaska. The data period is 1966 to 1998.

through the glacier (Mayo 1992). Geothermal heat flow is assumed to be $1.2 \mu\text{cal}/\text{cm}^2\text{s}$. The potential energy lost by ice motion is estimated by assuming a dynamic equilibrium in which the glacier annually transports the long-term average firm balance from the accumulation area to the ablation area. The altitude loss is assumed to be the difference of the area-weighted average altitude of the accumulation area and the area-weighted average altitude of the ablation area. The potential energy lost by water flowing through and under the glacier is estimated by assuming that 95% of the basin runoff descends from the ELA to the terminus. At Gulkana Glacier, the area-averaged melting caused by the three components for 1993 were evaluated by March and Trabant (1997): geothermal heat flow produced 0.005 m/year water equivalent; ice motion produced an estimated 0.005 m/year water equivalent; and flowing water produced 0.06 m/year water equivalent.

Errors. An unavoidable consequence of area integration is that the possibility of rigorous error analysis ends when area integration begins. Unassessable errors of extrapolation and area integration reduce confidence in the glacier and basin mass balances. There is seemingly no rigorous way to assess the error of the glacier-averaged or basin-averaged mass balance by analysis of the input data alone.

Assessment of the errors in the area-averaged

mass balances requires an independent determination of a related quantity. At one time it was hoped that the hydrologic balance would serve as an error checking – validation companion to mass-balance programs. However, it became apparent that large errors in both data sets preclude all but the most general mutual verification. Another possibility is to measure glacier volume changes over 5- or 10-year periods. Several methods are applicable. Carefully controlled photogrammetric analysis of large-scale stereo aerial photography, carefully controlled geodetic surveys, airborne laser-altimeter profiling, and satellite remote sensing of glaciers are each capable of determining volume changes of glaciers with sufficient accuracy so that cumulative errors in the surface mass balance data can be analyzed. In the near future, satellite remote sensing data are expected to provide seasonal assessments of glacier volume change.

Cumulative net mass balance bias and glacier volume changes

The difference between the cumulative net mass balance and measured volume change of Wolverine Glacier for an 11-year period suggests that there is a bias in the evaluation of net mass balance. Wolverine Glacier is the only glacier in Alaska for which a volume change has been used to evaluate cumulative biases in annual mass balances. Geodetic surveys on Wolverine Glacier during 1974

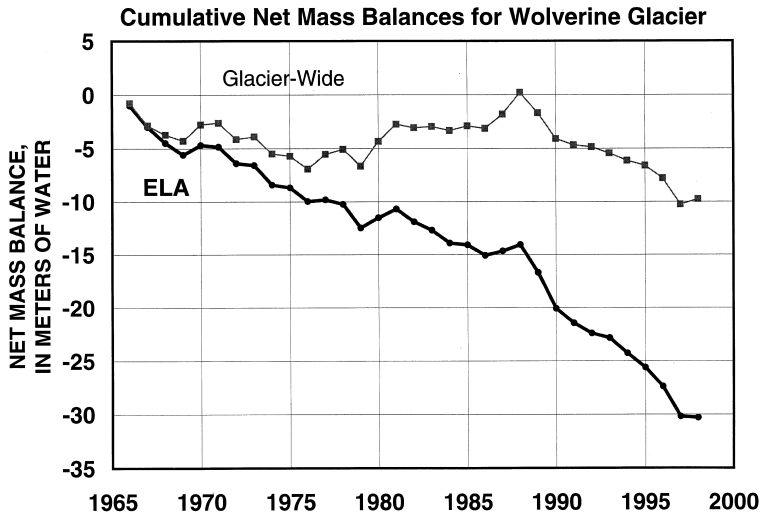


Fig. 4. Cumulative net mass balances measured at a site near the Equilibrium Line Altitude (ELA) of Wolverine Glacier and the glacier-wide mass balance for Wolverine Glacier. Note that there is no period of mass increase suggested by the measurement series made near the ELA. The measurements made near the ELA become more negative with time because the measurement site is a little below the current ELA. The data period is 1966 to 1998.

and 1985 were directly compared at 66 locations along one longitudinal and three transverse profiles. From these comparisons, a net volume increase of $55 \times 10^6 \text{ m}^3$ was estimated for the period (Mayo *et al.* 1985). The geodetically measured volume increase amounts to an average thickening of $3.0 \pm 0.7 \text{ m}$ (2.6 m water equivalent) for Wolverine Glacier. The cumulative net mass balance for the period was 7.7 m water equivalent. Therefore, an adjustment of -0.46 m/year was determined for the net mass balance for the period between 1974 to 1985. This adjustment makes the surface mass balance more negative, which helps reduce the widely observed difference between the hydrologic and surface mass balance.

When using measured volume changes to correct mass balance, consideration must be given to the fractional part of a year that separates two glacier-surface altitude determinations and changes in the average density of a glacier. The effects of a fractional part of a year that is part of a volume-change interval must be removed. The fraction of a year is important because the seasonal altitude variations caused by mass balance and glacier flow may be relatively large compared to the change during a period of years (Mayo and Trabant 1986). Furthermore, glaciers that are losing mass have higher bulk densities than glaciers that are increasing in volume (Krimmel 1989).

Measurements of the gradual volume change of glaciers are a reminder of the importance of regularly updating the area-altitude distribution used

for integrating measurements with area for glacier and basin values. Often, year-to-year changes are negligibly small and appropriately ignored. However, after 5 or 10 years of a continuous mass-balance trend, the cumulative effect must be evaluated and a correction retrofit into the interpreted data, thereby removing a bias in the reported values. Furthermore, because of analytical errors, meaningful changes in the volume, and therefore the area-altitude distribution, of a glacier may only be assessable after 5 or 10 years. This means that area-integrated mass balances must be revised retrospectively as changes in the area-altitude distribution are defined. In another approach, Krimmel (1996) has demonstrated the use of annual aerial photogrammetric analysis for area-altitude integration of mass balance measurements, thus avoiding the necessity of revising previously reported glacier and basin mass balances.

Mass-balance for climate correlations

If investigation of the relation between mass balance and climate is the principal goal, it is important that a long-term mass-balance data set contain as few random errors, biases, and hidden assumptions as possible. With this goal in mind, the original objectives of this section of the report were to (1) demonstrate the close correlation between net mass balance measured near the equilibrium line and the net mass balance of the glacier (Fig. 3); (2) suggest that the directly measured data are better

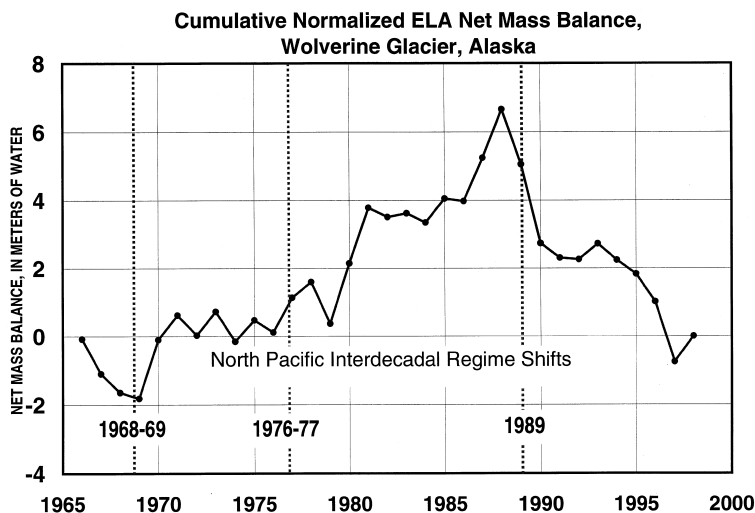


Fig. 5. Accumulated normalized net mass balances measured near the Equilibrium Line Altitude (ELA) of Wolverine Glacier. The data period is 1966 to 1998.

for correlation with climate variations than are the area-integrated values that incorporate possible bias and unassessable extrapolation errors; and (3) propose that the minimum mass-balance program needed for correlation with climate is an intensively measured site near the equilibrium line of a glacier somewhat similar to that proposed by Koerner (1986).

However, despite strong correlations between the annual net mass balances measured near the ELA and the glacier-averaged net mass balance (0.84 at Gulkana Glacier and 0.91 at Wolverine Glacier, Fig. 3), the cumulative net mass balance for the sites near Wolverine Glacier's ELA does not show the 1977–88 growth period (Fig. 4). The overall negative trend of the measurements is because the measurement site 'near' the ELA is actually in the upper part of the ablation area. The trend was removed from the 'near ELA' data by normalizing the measurements, and the normalized series was accumulated and plotted (Fig. 5). The graph (Fig. 5) of the cumulative normalized data suggests that Wolverine Glacier was growing during the period 1968 to 1988. Currently, no volume-change measurements confirm or deny glacier volume increase during the 1968 to 1977 period. However, measurements and observations in the basin agree with growth only after 1977. All three years (1968, 1977, and 1988) are recognized transitions in the climate of the north Pacific basin (see Fig. 2), specifically the Aleutian Low and Arctic Oscillation climate indices (Overland

et al. 1999). The period of geodetically confirmed growth of Wolverine Glacier occurred between the 1976–77 and 1989 decadal shifts. During that period, the Aleutian Low index reached the extreme low values for the 20th century (Overland *et al.* 1999, fig. 1a). During the 1968 to 1977 period, the Aleutian Low index was decidedly in the opposite state (the atmospheric low-pressure centers were at anomalously high latitudes), which should not have produced growth of Wolverine Glacier.

It is also increasingly important that seasonal mass balances, as opposed to 'annual' values only, be measured and reported and, that the evolution of data collection and analysis methods not destroy the continuity of the results. Seasonal components of glacier mass balance are fundamentally important for correlations with climate indices (e.g. Hodge *et al.* 1998). Long-term mass-balance programs inevitably evolve in both field and analytical methods. As these changes occur, retrospective re-evaluation of previously reported results should be expected and accepted by the community of data users.

Conclusions

The conclusion is that attempting climate correlations with a time-series of measurements made 'near' the ELA, whether or not a trend has been removed, is not advisable. Furthermore, it is likely that three measurement sites – one low in the abla-

tion area, one near the ELA, and one in the accumulation area – is the minimum number of measurement points, even for relatively small glaciers.

Furthermore, measuring glacier responses to climate-induced mass-balance changes requires a commitment to long-term programs that measure seasonal and annual mass balances. Glacier-climate programs must (1) be enduringly funded; (2) regularly update the area–altitude distributions used for the area integration of the balance measurements at sites; (3) be carefully checked for the effects of cumulative bias errors; and (4) ensure that changes in analytical techniques do not disrupt the continuity of the data set. Enduring funding may always be problematic. Updating the area–altitude distribution by remapping from aerial or satellite image sources also serves as a check on cumulative errors and changing analytical methods. Remapping enables evaluation of the net volume change of a glacier. If the measured volume change, carefully corrected for seasonal and bulk density differences, is equal to the cumulative mass balance for the same period, the suggestion is that the measured mass balances have no important bias errors. However, good agreement says little about the magnitude of randomly distributed errors.

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Manuscript received August 1998, revised and accepted November 1998.